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# Major optical depth perturbations to the stratosphere from volcanic eruptions: Stellar extinction period, 1961–1978

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**Abstract.** A revised chronology of stratospheric aerosol extinction due to volcanic eruptions has been assembled for the period 1961–1978, which immediately precedes the era of dedicated satellite measurements. On the whole, the most accurate data consist of published observations of stellar extinction, supplemented in part by other kinds of observational data. The period covered encompasses the important eruptions of Agung (1963) and Fuego (1974), whose dust veils are discussed with respect to their transport, decay, and total mass. The effective (area-weighted mean) radii of the aerosols for both eruptions are found to be 0.3–0.4 $\mu$m. It is confirmed that, among known tropical eruptions, Agung's dust was unique for a low-latitude eruption in remaining almost entirely confined to the hemisphere of its production. A new table of homogeneous visual optical depth perturbations, listed by year and by hemisphere, is provided for the whole period 1881–1978, including the pyrheliometric period before 1961 that was investigated previously.

## 1. Introduction

New estimates of major optical depth perturbations to the stratosphere during the period 1881–1960 were given in a recent paper [*Stothers*, 1996]. All of the turbidity anomalies investigated there arose from large volcanic eruptions, which raise sulfur gases aloft to form sulfuric acid aerosols in the stratosphere. Values of the optical depth perturbations in each case were estimated chiefly by using published pyrheliometry and, to a lesser extent, other published data such as polarimetry of blue sky light. To carry the tabulation up to the start of the modern era of dedicated satellite measurements, which began in November 1978 with the activation of the Stratospheric Aerosol Measurement II (SAM II) instrument [*McCormick et al*., 1979], the present paper utilizes various kinds of ground-based data for the years 1961–1978. Owing to improved accuracy of photoelectric stellar extinction measurements, which, at good observing sites, yield results superior to most pyrheliometric extinction measurements, our analysis relies primarily on data acquired in the course of nighttime observations of stars, at various zenith angles, from astronomical observatories both north and south of the equator.

*Sato et al*. [1993] have previously studied the period 1961–1978. They have tabulated pyrheliometric data compiled by *Dyer and Hicks* [1968] for the period 1961–1965 as well as visual lunar eclipse data obtained by *Keen* [1983] for the period 1960–1982. As we shall see, however, nearly all of these data are too coarse and, in the case of the lunar eclipse data, too sparse to provide anything but a first reconnaissance, however useful, of the present problem. In addition, Sato et al. have equated the pyrheliometric optical depth to the visual optical depth, whereas the two are now known to differ by a factor of about 60% in the case of small volcanic aerosols [*Stothers*, 1997].

In section 2 we describe the methodology and principal data used for the present study. Conversion formulae to obtain the visual optical depth perturbation from other extinction measures are presented in section 3. The temporal developments of the various stratospheric dust veils produced by volcanic eruptions during the period 1961–1978 are then discussed in section 4. Section 5 contains new calculations of the effective particle sizes for the two largest dust veils, whose total masses are finally derived in section 6. The general discussion of section 7 includes a table and a graphical representation of the annual mean visual optical depth perturbations by hemisphere for the extended period 1881–1978.



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## 2. Methodology and Principal Data Used

The methodology employed in this paper follows the procedures described in our 1996 paper, but is briefly reviewed here. Measurements of atmospheric attenuation of starlight or of direct sunlight (referred to the local zenith) are taken from the published literature. To determine the possible departures from average conditions at any site, the "reference year" method is used, in which monthly mean extinction coefficients for each year are calculated and then examined to locate periods of very low extinction, often covering several years. The turbidities in these relatively unperturbed years are next averaged, month by month. A simple subtraction of each monthly mean extinction coefficient from its average value under unperturbed conditions yields the extinction perturbation for that month. In cases where only the annual mean extinction coefficients have been published (or can be considered reliable), a similar procedure has been followed by using yearly means instead of monthly means. However, the yearly means are always taken to be averages of the monthly means.

In general, monthly averages have been computed for all the data sets used here, because seasonal variations in the amounts of ozone, water vapor, and tropospheric particulates induce significant annual cycles in the background extinction at some sites, especially in the visual band, which is of greatest importance here. There also exist, at any given site, small year-to-year fluctuations in the monthly averages that are not obviously

**Table 1.** Stations With Stellar Extinction Data Subjected to New Reduction

| Station | Country | Latitude | Longitude | Period | Sources of Data |
|---|---|---|---|---|---|
| Bloemfontein | South Africa | 29°S | 26°E | 1963–1965 | *Irvine and Peterson* [1970] |
| Cerro Tololo | Chile | 30°S | 71°W | 1961–1980 | *Moreno and Stock* [1964], *Moreno et al*. [1965], *Gutiérrez-Moreno et al*. [1967], *Stock* [1969], *Gutiérrez-Moreno and Moreno* [1970], *Moreno and Maza* [1974], *Gutiérrez-Moreno et al*. [1982] |
| Flagstaff, Arizona | United States | 35°N | 112°W | 1955–1985 | *Jerzykiewicz and Serkowski* [1966], *Jerzykiewicz* [1972], *Lockwood and Thompson* [1986] |
| Kitt Peak, Arizona | United States | 32°N | 112°W | 1960–1969 | *Lockwood and Hartmann* [1970] |
| La Silla | Chile | 29°S | 71°W | 1971–1985 | *Groenbech et al*. [1976], *Sterken and Jerzykiewicz* [1977], *Rufener* [1986] |
| Mount Bingar | Australia | 34°S | 146°E | 1962–1964 | *Hogg* [1963], *Przybylski* [1964] |
| Mount Locke, Texas | United States | 31°N | 104°W | 1960–1980 | *de Vaucouleurs* [1965], *Neff and Travis* [1967], *de Vaucouleurs and Angione* [1974], *Angione and de Vaucouleurs* [1986] |
| Pretoria | South Africa | 26°S | 28°E | 1962–1964 | *Hill* [1964] |
| San Pedro Mártir | Mexico | 31°N | 115°W | 1973–1979 | *Schuster* [1982] |

attributable to known volcanic eruptions [*Roosen et al*., 1973; *Laulainen et al*., 1977]. It is important to keep this fluctuating background extinction in mind when looking for small volcanically induced perturbations to the stratosphere.

For the purposes of this paper, we assume that volcanically undisturbed periods are represented by the intervals 1949–1963 (up to March 1963) and 1977–1979. In order to use the 1963–1965 Bloemfontein observations [*Irvine and Peterson*, 1970], we treat 1965 as having been minimally disturbed. Finally, we completely ignore the apparently anomalous 1960 Kitt Peak data [*Lockwood and Hartmann*, 1970].

The stations at which most of the data of this paper were acquired are listed in Table 1 for the stellar extinction measurements and in Table 2 for the pyrheliometric measurements. The time periods covered and the references for the data used are also tabulated. Notice the midlatitude concentrations of the stations in both hemispheres: all of the listed stations lie between 20° and 40°, with the exception of the McMurdo and South Pole stations in Antarctica. (For convenience, we shall henceforth consider the McMurdo observations as belonging to the South Pole group.) Data for many other stations have been examined, but are found to be generally too noisy for our purposes. References to such data, however, have been provided by *Laulainen and Hodge* [1972], *Taylor et al*. [1977], *Laulainen* [1977], *Dyer and Hicks* [1968], and *Dyer* [1974].

Other methods of obtaining visual optical depth estimates are available. During a total eclipse of the Moon, sunlight passing through the Earth's atmosphere is refracted and scattered into the shadow cone by the normal atmospheric molecules but tends to be blocked by the transient larger particles, like those created in a volcanic eruption. *Keen*'s [1983] study of the darkness of total lunar eclipses in the period 1960–1982 provides a very homogenous, though sparse, database of stratospheric optical depths. Although there are dangers with this method, Keen's results agree well with, and may have been partly calibrated by, the more fundamentally derived values that were discussed by *Matsushima* [1967] for the eclipses of December 30, 1963 [*Hansen and Matsushima*, 1966], June 25, 1964 [*Bouska and Mayer*, 1965], and December 19, 1964 [*Matsushima et al*., 1966]. The disadvantages of the lunar eclipse method are formidable, however: on the average, only one

**Table 2.** Stations With Pyrheliometric Data Subjected to New Reduction

| Station | Country | Latitude | Longitude | Period | Sources of Data |
|---|---|---|---|---|---|
| Aspendale | Australia | 38°S | 144°E | 1959–1972 | *Dyer and Hicks* [1965], *Dyer* [1974] |
| Mauna Loa, Hawaii | United States | 20°N | 156°W | 1958–1977 | *Ellis and Pueschel* [1971], *Pueschel et al*. [1972], *Mendonca et al*. [1978], *Shaw* [1979][a] |
| McMurdo Station | Antarctica | 78°S | 161°E | 1949–1966 | *Fischer* [1967][a] |
| South Pole | Antarctica | 90°S | | 1957–1966 | *Flowers and Viebrock* [1965], *Viebrock and Flowers* [1968] |
| Tucson, Arizona | United States | 32°N | 111°W | 1956–1983 | *Heidel* [1972], *Szymber and Sellers* [1985] |

[a]Sunphotometer measurements at $\lambda = 0.5$ $\mu$m.

night a year has a total lunar eclipse; geographic resolution is essentially hemispheric in scale; and tropospheric clouds along the terminator may be contributing to the measured turbidity.

Optical probing of the stratosphere by searchlight and by lidar (laser radar) can also be used to derive the visual optical thickness, but the values so derived often have large uncertainties because they are not direct measures. Such probing with beams, however, provides useful information on the height and time development of the scattering layers.

The only other method that will be used here is based on the measured percentage of blue sky polarization. This is discussed in the next section.

## 3. Conversion Formulae

Conversion formulae needed for the reduction of data in the present paper are as follows. The optical depth $\tau_\lambda$ at any wavelength $\lambda$ is related to the astronomical extinction $a_\lambda$, measured in stellar magnitudes, by [*Hardie*, 1962]

$$\tau_\lambda \approx 0.921 a_\lambda. \tag{1}$$

If the pyrheliometrically measured extinction is expressed as the Linke turbidity factor $T$ for solar radiation of all wavelengths passing through one air mass, then [*Szymber and Sellers*, 1985]

$$\tau_{\mathrm{pyr}} \approx 0.08 T. \tag{2}$$

In the case of small sulfuric acid aerosols that have been produced by volcanic eruptions, the relation between the pyrheliometric extinction and the visual ($\lambda = 0.55$ $\mu$m) extinction is given by

$$\tau_{\mathrm{vis}} \approx 1.6 \tau_{\mathrm{pyr}}, \tag{3}$$

owing to the fact that the effective wavelength of pyrheliometers for such small aerosols lies close to 0.8–0.9 $\mu$m while $\tau_\lambda \propto \lambda^{-1}$, approximately [*Volz*, 1970a, 1975a; *Stothers*, 1996, 1997; *Molineaux et al.*, 1998]. Simultaneous measurements of atmospheric transmission with pyrheliometers and with spectrobolometers following the eruptions of Santa Maria (1902), Ksudach (1907), and Katmai (1912) that were made by members of the Astrophysical Observatory of the Smithsonian Institution confirm, unambiguously, that $\tau_{\mathrm{vis}}/\tau_{\mathrm{pyr}} = 1.56 \pm 0.04$ [*Stothers*, 1997].

After the Agung (1963) eruption, however, the aerosols that drifted to the South Pole apparently acquired, from the low temperatures, an untypical size distribution, which *Volz* [1970a] inferred would lead to $\tau_{\mathrm{vis}}/\tau_{\mathrm{pyr}} \approx 1.2$. With some uncertainty, therefore, we adopt this smaller ratio for the 1963–1966 South Pole pyrheliometric observations [*Flowers and Viebrock*, 1965; *Viebrock and Flowers*, 1968].

Blue sky polarization measurements have also sometimes been made. If $p_N$ and $p_D$ represent the observed percentages of polarization during normal and disturbed periods, then

$$\tau_{\mathrm{pyr}} \approx 0.21 \ln (p_N/p_D), \tag{4}$$

which is based on data acquired both before and after the eruptions of Santa Maria (1902), Ksudach (1907), and Katmai (1912) [*Stothers*, 1996].

## 4. Volcanic Eruptions

### 4.1. Agung 1963

Agung Volcano (8°S), located on the island of Bali in Indonesia, erupted paroxysmically on March 17, 1963. Some of the eruption products penetrated the lower stratosphere, leading to vivid twilight glows in Indonesia and in north Australia during late March [*Weinert*, 1967]. By April the enhanced glows were noticed also in south Australia [*Hogg*, 1963; *Przybylski*, 1964; *Weinert*, 1967] and South Africa [*Hill*, 1964; *Burdecki*, 1964]. Appearing somewhat later in the northern hemisphere, the glows were first detected during May from the southwest United States [*de Vaucouleurs*, 1965] as well as from Europe [*Volz*, 1965]. The colors intensified in September [*Meinel and Meinel*, 1963; *Volz*, 1964] and even more so during the following few months [*Meinel and Meinel*, 1964, 1967; *Lamb*, 1966; *Volz*, 1969a]. This characteristic pattern followed closely what had been seen in Europe after the great eruption of Tambora (8°S), Indonesia, in April 1815 [*Stothers*, 1984].

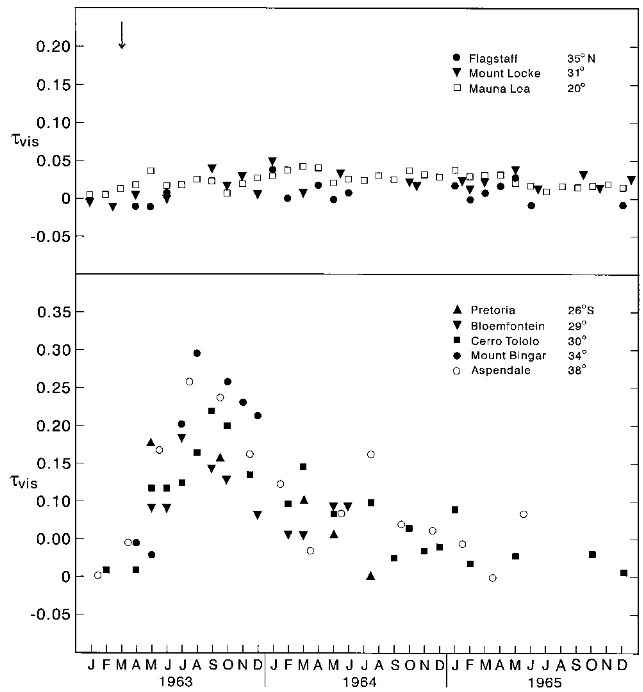


**Figure 1.** Visual optical depth perturbation due to the eruption of Agung (March 1963). The arrow indicates the date of the eruption.

Bishop's rings were detected around the Sun for about a year, between the summer of 1963 and the following summer [*Volz*, 1970a]. In April 1964 the outer radius of the ring was measured as being 20° [*Bullrich et al.*, 1968], which agrees well with measurements made after many other large volcanic eruptions [*Stothers*, 1996].

Stellar extinction and solar extinction data, reduced as outlined in sections 2 and 3, are plotted against time in Figure 1. In the southern hemisphere the visual turbidity rose to a sharp maximum in August and September 1963, reaching a value of 0.24. This peak can also be seen, more approximately, in published plots of less accurate data for several sites that are not included here: Kenya (1°S) and Congo (4°S) [*Dyer and Hicks*, 1968] and Zimbabwe (20°S) [*Lamb*, 1970].

Over the South Pole the turbidity peak appears to be shifted in time to January 1964 [*Flowers and Viebrock*, 1965; *Viebrock and Flowers*, 1968]. This is illustrated here in Figure 2, which compares South Pole data with the average turbidity at southern midlatitudes. The substantial delay occurs because the southward meridional transport of air is blocked by the strong

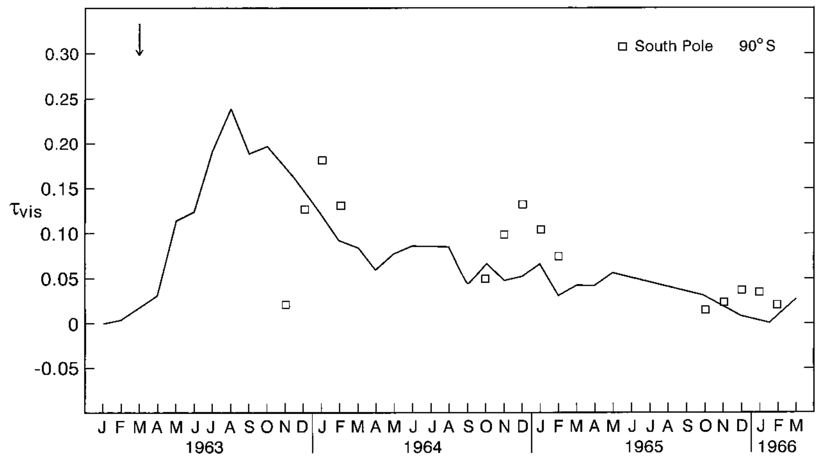


**Figure 2.** Visual optical depth perturbation at the South Pole due to the eruption of Agung (March 1963). A mean curve for 26°–38°S is also shown. The arrow indicates the date of the eruption.

polar vortex during the austral winter; after the polar vortex breaks down in the spring, meridional transport to the pole resumes. The fast buildup of a turbidity peak during the late spring and early summer repeats itself in 1965 and in 1966.

One need not assume that these annual waves of highly turbid air drifting into the south polar region are due to anything much more than the effects of the polar vortex. However, *Dyer and Hicks* [1965, 1968] and *Dyer* [1970, 1971] have argued that there exists a large-amplitude annual oscillation in the outflow of aerosols from an equatorial stratospheric reservoir. Their evidence for this oscillation is the apparent winter peaking of the 1964–1966 pyrheliometric optical depths over Aspendale, Australia, a feature that they have also identified in pyrheliometric data (of poorer quality) from elsewhere. On the other hand, the more accurate stellar extinction data shown in Figure 1 suggest that the winter peaking is less marked than was originally claimed by Dyer and Hicks. A very large peaking has been challenged, as well, by *Volz* [1970a, 1971] and by *Pueschel and Ellis* [1972] on similar grounds. Evidence from volcanic eruptions during the recent satellite period, however, has supported the notion of a tropical aerosol reservoir, from which there occurs poleward transport, modulated somewhat by the seasonal and quasi-biennial changes of the stratospheric meridional wind circulation patterns [e.g., *Trepte and Hitchman*, 1992; *Hitchman et al.*, 1994]. Since the transport and consequent mixing of the volcanic aerosols is more rapid in the lower layers than in the less populated upper layers of the stratosphere, any variations arising in the outflow from the tropical reservoir will become much less noticeable as soon as the stratospheric aerosol loading is approximately hemispherically uniform [*Barnes and Hofmann*, 1997]. This expectation may account for the rather muted seasonal amplitudes appearing in the stellar extinction data in Figure 1.

Turbidity over the northern hemisphere in 1963 barely nudges above the noise level. It peaks at about 0.03. *Volz* [1970a] showed a somewhat noisier plot because he did not subtract a prevolcanic background from most of the data that he used and because he employed many records that are too short to be accurately calibrated. *Dyer and Hicks* [1968] extracted a considerably higher northern turbidity, chiefly because they included a great many records from polluted atmospheric environments in Japan, the United States, and the Soviet Union. Agung's true signal is almost impossible to pick out in such turbid sites [*Flowers et al.*, 1969; *Pivovarova*, 1970; *Clemesha*, 1971; *Volz*, 1971; *Dyer*, 1974; *Yamauchi*, 1995] or, at best, its signal is very hard to quantify accurately [*Volz*, 1969b; *Hoyt et al.*, 1980; *Hoyt and Fröhlich*, 1983]. Searchlight and lidar techniques, however, confirm that the true excess turbidity over the United States was roughly 0.01–0.03 in mid-1963 and roughly 0.01–0.02 in 1964 [*Grams and Fiocco*, 1967; *Elterman et al.*, 1973].

Although even the best northern data are too uncertain to evaluate the stratospheric residence time of Agung's dust, the dust in the southern hemisphere is found to have decayed approximately exponentially, with an $e$-folding time of 1.0 ± 0.1 years. This decay is due to aerosol sedimentation and to stratosphere-troposphere air mass exchanges. Table 3 lists the values of the monthly mean visual optical depths in each hemisphere, from the fast rise of turbidity in April 1963 to the slow decay beginning in August and September 1963 and continuing on to May 1965. The mass loading ratio between the southern and northern hemispheres, according to our and *Volz*'s [1970a]

**Table 3.** Visual Optical Depth Perturbations (Multiplied by 1000) Due to the Agung Eruption of March 1963

| Year | Latitude Range | Jan. | Feb. | March | April | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1963 | −40° to −20° | | | | 33 | 116 | 126 | 192 | 239 | 190 | 196 | 176 | 148 |
| | +20° to +40° | | | | 5 | 17 | 10 | 18 | 24 | 31 | 12 | 25 | 15 |
| 1964 | −40° to −20° | 123 | 92 | 84 | 60 | 79 | 88 | 87 | 87 | 47 | 68 | 49 | 51 |
| | +20° to +40° | 39 | 19 | 24 | 29 | 18 | 23 | 24 | 29 | 24 | 29 | 25 | 30 |
| 1965 | −40° to −20° | 67 | 31 | 43 | 43 | 57 | | | | | | | |
| | +20° to +40° | 27 | 16 | 21 | 24 | 30 | | | | | | | |

assessment of the data, is approximately 8:1. In situ particle sampling, after subtraction of the prevolcanic background, suggests very nearly the same ratio [*Castleman et al.*, 1974]. These findings cast doubt on the 3:1 ratio that *Dyer and Hicks*'s [1968] data imply.

Indirect evidence of the magnitude of the optical depth perturbations in 1963–1965 can also be derived by using various other methods. First of all, there are the twilight glows, which in South Africa and Australia were reported as having been most intense during April and May 1963 [*Burdecki*, 1964; *Weinert*, 1967] and at Cerro Tololo, Chile, as having continued into July (J. Stock in the work of *Hogg* [1963]). From data compiled after early twentieth century volcanic eruptions [*Stothers*, 1996], we would expect to find a rapid suppression of the glows starting when $\tau_{vis}$ exceeded 0.15. According to Figure 1, Agung twilights indeed began to weaken when this value of $\tau_{vis}$ had been exceeded.

A second method of inferring turbidity is to measure the amount of blue sky polarization. In India, on January 11 and February 1, 1964, the polarization was 47%, down from a probably normal 62% on May 30, 1955 [*Shah*, 1969]. Using (4), this decrease seems to imply $\tau_{vis} = 0.09$. Similarly, the percentage of blue sky polarization measured in Hawaii on April 10–11, 1964, and on September 5, 1965, was 60% and 75%, respectively [*Bullrich et al.*, 1968] ostensibly implying $\tau_{vis} = 0.07$. These turbidity estimates, however, are exaggerated values because the examples given refer to hazy days when the polarization was notably reduced. Indeed, Figure 1 indicates that the mean value of $\tau_{vis}$ at northern midlatitudes during early 1964 was about 0.03. Nevertheless, the derived polarization results confirm that the turbidity in the northern hemisphere was much smaller than was the case in the south.

Observations of total eclipses of the Moon provide a third way of estimating turbidity. Measured visual optical depth perturbations of 0.13 on December 30, 1963 (southern hemisphere), 0.08 on June 25, 1964 (southern hemisphere), and 0.05 on December 19, 1964 (northern hemisphere) [*Matsushima*, 1967; *Keen*, 1983], although very rough values, match in a general way the stellar extinction data shown in Figure 1.

### 4.2. Fuego 1974

Fuego Volcano (14°N) in western Guatemala erupted on October 14 and 17, 1974, and spewed aerosol-forming gases into the lower stratosphere. The twilight glows, originating just above the tropopause, were detected not long after at northern midlatitudes [*Meinel and Meinel*, 1975; *Volz*, 1975b; *Shaw*, 1975] and even as far north as College (65°N), Alaska [*Shaw*, 1975]. Bishop's ring appeared around the Sun in November 1974 with a measured outer radius of about 27° [*Meinel and Meinel*, 1975]. The rapidly moving, small-scale irregularities of the dry haze in the stratosphere apparently caused the astronomical seeing (a measure of the angular sizes of stars) to deteriorate badly during 1975 as compared to neighboring years [*Abt*, 1980; *Walker*, 1984]. It is interesting to note that the same phenomenon had occurred after the eruptions of Tambora (1815) [*Stothers*, 1984], Krakatau (1883) [*Symons*, 1888, p. 225], and, apparently, Agung (1963) [*Walker*, 1970], while around the time of the mystery eruption of 536 [*Stothers*, 1996] the stars were described as "dancing in a strange manner" [*Hamilton and Brooks*, 1899, p. 231]. In September 1975 the northern midlatitude haze was still seen, though much more weakly [*Scorer*, 1975; *Volz and Shettle*, 1976], and it essentially disappeared after the middle of 1977 [*Volz*, 1981].

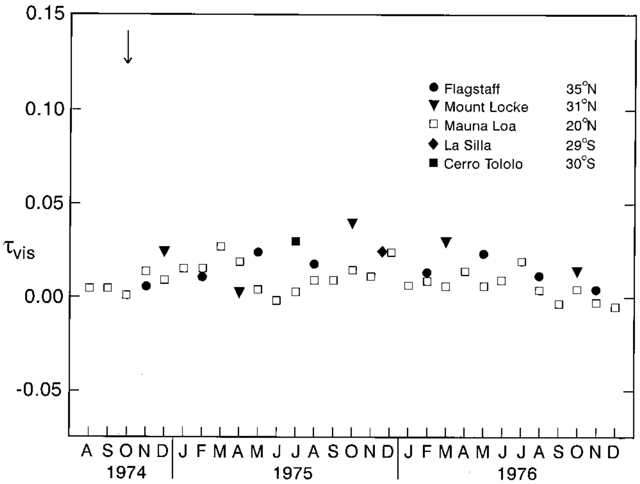


**Figure 3.** Visual optical depth perturbation due to the eruption of Fuego (October 1974). The arrow indicates the date of the eruption.

The detailed time development of the visual atmospheric extinction created by the Fuego eruption is plotted in Figure 3. The rather confusing scatter of measurements is greater than what one might have expected for data obtained at such optimally chosen observing sites. Close examination reveals that most of the scatter arises from the unexpectedly low pyrheliometric turbidities that were measured at Mauna Loa after March 1975. Since instrumental calibration errors have been such a persistent problem at Mauna Loa that special data reduction methods have been required to minimize them [*Mendonca et al.*, 1978], it may be legitimate to question the accuracy of the Mauna Loa results. Ignoring them, we find that the turbidity reaches a maximum of about 0.03 in the middle to late months of 1975 and then falls off roughly exponentially, with an *e*-folding time of 1.0 ± 0.4 years. A range of *e*-folding times from 0.6 to 1.0 years has previously been found for this eruption by means of lidar observations [*Russell and Hake*, 1977; *McCormick et al.*, 1978; *Swissler et al.*, 1982] and in situ particle sampling [*Hofmann and Rosen*, 1977, 1987; *Lazrus et al.*, 1979; *Sedlacek et al.*, 1983]. Lidar observations have also yielded 0.025 as an average excess turbidity for the whole year 1975 [*Russell and Hake*, 1977; *Itabe et al.*, 1977], although *Keen* [1983] has estimated an excess turbidity of about 0.04 from observations of the total lunar eclipse of May 25, 1975.

All the evidence of Figure 3 suggests that Fuego's aerosols eventually spread about equally in both hemispheres. Lidar data confirm this result [*Remsberg et al.*, 1982].

### 4.3. Other Eruptions

Lists of explosive volcanic eruptions that possibly disturbed the stratosphere during the years 1961–1978 have been published by many authors [e.g., *Lamb*, 1970; *Cronin*, 1971; *Castleman et al.*, 1974; *Mendonca et al.*, 1978; *Newhall and Self*, 1982; *Sedlacek et al.*, 1983]. In addition to Agung (1963) and Fuego (1974), the likeliest perturbers of the stratosphere were five other tropical volcanoes, Taal (1965), Kelut (1966), Oldoinyo Lengai (1966), Awu (1966), and Fernandina (1968).

Can the dates of these lesser eruptions be reconciled with the detailed record of observed stratospheric extinction? Annual mean visual optical depth perturbations are shown for each year from 1961 to 1978 in Figure 4.

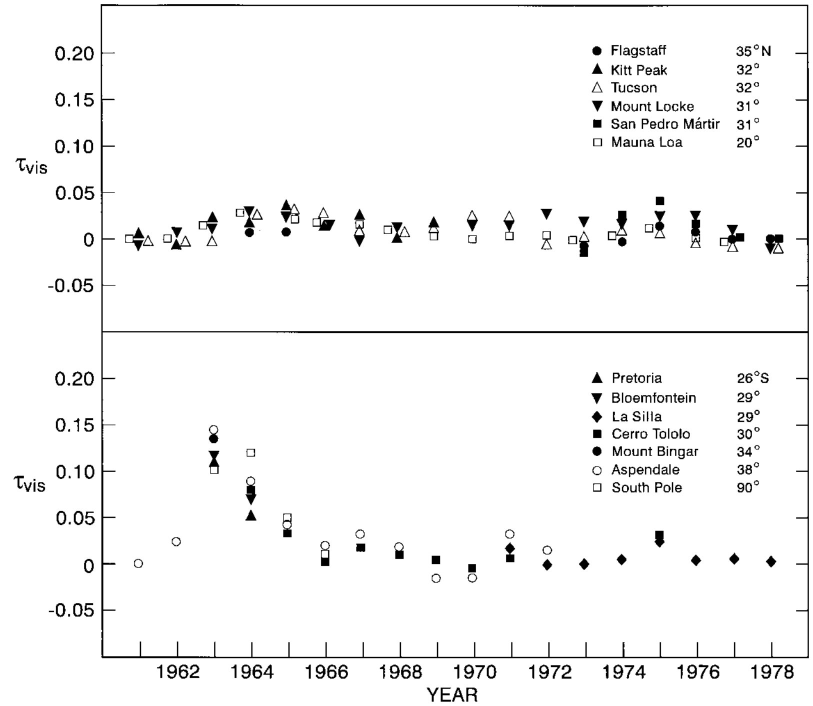


**Figure 4.** Visual optical depth perturbation by year, 1961–1978.

Most notable is the fact that in the northern hemisphere the post-Agung turbidity does not decline rapidly after 1966, but remains elevated, though at a low level, through 1971. This is confirmed by twilight observations [*Volz*, 1969a, 1970b, 1974] and also by lunar eclipse observations [*Volz*, 1970a; *Keen*, 1983]. In the southern hemisphere a slight post-Agung turbidity enhancement occurs in 1967 (probably due to the four 1965–1966 tropical eruptions) with another increase coming in 1971. Independent measurement techniques confirm these observations: specifically, lidar does [*Clemesha and Nakamura*, 1972; *Russell and Hake*, 1977] and also searchlight probing [*Elterman et al.*, 1973] and in situ particle sampling [*Castleman et al.*, 1974; *Lazrus and Gandrud*, 1974; *Bigg*, 1976]. The source of the late 1971 turbidity increase in the southern hemisphere is uncertain, but *Dyer* [1974] has suggested the August 1971 eruption of Mount Hudson (46°S), Chile. An unrelated dust event that occurred in October 1971 in the northern hemisphere might have been due to an unidentified North Pacific volcano [*Volz*, 1974] or to the small September 1971 eruption of Fuego [*Hofmann et al.*, 1975].

A total lunar eclipse on October 6, 1968, was dark enough to indicate $\tau_{vis} = 0.06$ [*Keen*, 1983]. The unusually high turbidity that night was attributed by Keen to the eruption of Fernandina (0°S) in the Galapagos Islands the preceding June. According to Figure 4, however, stellar extinction measurements indicate a very low turbidity of 0.01 for 1968, a year when the astronomical seeing was also quite normal [*Walker*, 1971].

*Shaw*'s [1979] observation of a small optical depth perturbation between March and August 1976 over Mauna Loa, Hawaii, which he considered to have been possibly due to an eruption of Augustine (59°N), Alaska, the preceding January, is more likely the tail end of Fuego's decaying turbidity. The Augustine eruption column was almost certainly too low to have reached the tropopause [*Meinel et al.*, 1976; *Meinel and Meinel*, 1983].

## 5. Effective Particle Sizes

### 5.1. Agung Aerosols

The wavelength dependence measured for an optical depth perturbation supplies information about the properties of the particles that are present in the disturbing layer. During 1963 and 1964, stellar photometry in the southern hemisphere in several wavelength bands ranging from 0.32 $\mu$m to 0.59 $\mu$m yielded a nearly flat profile of optical extinction [*Hogg*, 1963; *Przybylski*, 1964; *Hill*, 1964; *Moreno et al.*, 1965; *Irvine and Peterson*, 1970]. This result indicates that the particle sizes were significantly greater than molecular. To constrain the particle sizes more closely, longer wavelengths are needed. B. Westerlund (in the work of *Hogg* [1963]) secured a few observations at 0.67 and 0.80 $\mu$m, but these observations cannot be meaningfully combined with the shorter-wavelength observations of Hogg at the same observatory because the two sets were obtained a month apart, when the optical thickness was rapidly increasing. *Irvine and Peterson* [1970], on the other hand, acquired simultaneous observations of stellar extinction in South Africa over a broad wavelength range 0.36–1.06 $\mu$m, enabling both *Volz* [1970a] and *Deirmendjian* [1973] to plot sample spectral extinction curves for several selected dates.

A large monthly (and even daily) variability, however, occurs in the optical thickness. For this reason, we prefer to combine all the available South African data for 1963 into a single annual mean. On the basis of a total of 37 nights during the months of May, June, July, September, and December 1963, the resulting spectral extinction curve is shown in Figure 5.

To match the plotted data with a theoretical curve, we make the following assumptions about the aerosol properties, based on both observations and experiments conducted in the wake of the eruptions of Agung (1963) and of Pinatubo (1991): a spherical particle shape and a lognormal distribution of radii [*Mossop*, 1964], a composition of 75% sulfuric acid and 25% water [*Rosen*, 1971], a real part of the complex index of refraction, $\mathrm{Re}(m) = 1.43$ [*Baumgardner et al.*, 1996], and an imaginary part, $\mathrm{Im}(m) = 0$ [*DeLuisi and Herman*, 1977]. The vanishing imaginary part indicates that the particles are pure scatterers with no absorption component. Using Mie scattering theory, we are able to construct theoretical extinction curves, match them to the observed curve, and select the best fit [*Stothers*, 1997]. [In the latter paper a misprint occurs in equation (A7) for the theoretical $\tau_\lambda$: $(2\pi/\lambda)^2$ for $(2\pi/\lambda)^i$.] Numerous tests have previously shown that the effective (area-weighted) particle radius $r_{\mathrm{eff}}$ is not very sensitive to existing uncertainties in the particle composition, refractive index, and size distribution.

For Agung the best fit occurs when $r_{\mathrm{eff}} = 0.35$ $\mu$m. This case is shown in Figure 5, where the residuals around the fitted curve fall well below the estimated errors of the observational data. If the ultraviolet ($\lambda < 0.4$ $\mu$m) extinction measurements are ignored as being possibly less reliable than the others [*Taylor et al.*, 1977], $r_{\mathrm{eff}}$ still turns out to be 0.35 $\mu$m. On the basis of cruder analyses of the Agung turbidity data, the radius mode (which should be close to the effective radius for a small-variance lognormal distribution) was previously found to be ~0.3 $\mu$m by *Volz* [1970a] and ~0.5 $\mu$m by *Neumann* [1974], but only ~0.1 $\mu$m by *Deirmendjian* [1973], who used a somewhat inappropriate water-haze model. *Mossop*'s [1964] in situ collections of stratospheric aerosols between June and September 1963 indicate a time-average median radius of ~0.33 $\mu$m.

With less certainty, we may similarly analyze the South African extinction data for 11 nights in May 1964, which is all that can be reliably calibrated for that year. The May 1964 average extinction (which is not plotted) yields $r_{\mathrm{eff}} = 0.30$ $\mu$m.

### 5.2. Fuego Aerosols

*Schuster* [1982] has made multicolor ($\lambda = 0.33$–$1.10$ $\mu$m) stellar extinction measurements in Baja California, Mexico, during the 7 years 1973–1979 and has published these as annual means. To determine as accurately as possible the atmospheric optical depth perturbation in 1975, we average his results for the 6 other years and subtract this average as the unperturbed background. On the basis of seven nights in 1975 the resulting spectral plot is shown in Figure 5. Inversion of these data yields $r_{\mathrm{eff}} = 0.32$ $\mu$m. If the ultraviolet extinction measurements are ignored, $r_{\mathrm{eff}}$ increases slightly to 0.39 $\mu$m.

## 6. Masses of the Dust Veils

Adopting standard assumptions for the aerosol microphysical properties and also assuming longitudinal symmetry of the aerosol distribution over the globe, the total mass of the dust veil in the stratosphere in megatons (Tg) is given by

$$M_D = 75\ [\bar{\tau}_{\mathrm{vis}}(\mathrm{NH}) + \bar{\tau}_{\mathrm{vis}}(\mathrm{SH})], \qquad (5)$$

where an overbar denotes a hemispheric mean [*Stothers*, 1996]. The numerical coefficient has an estimated uncertainty of less than 20%. To determine $M_D$, $\bar{\tau}_{\mathrm{vis}}$ at the time of maximum turbidity should be used.

For Agung (1963) we obtain in this way a total aerosol mass

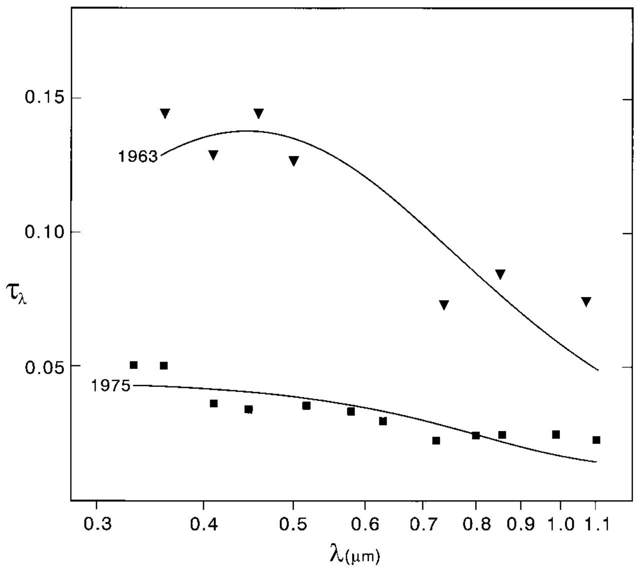


**Figure 5.** Theoretical spectral extinction curves fitted to the stratospheric optical depth perturbations detected over Bloemfontein, South Africa, in 1963 and over San Pedro Mártir, Mexico, in 1975.

of 20 Tg. This agrees fairly well with *Deirmendjian*'s [1973] cruder estimate of 15 Tg, once his value of 9 Tg for a normalized particle density of $\rho = 1.0$ g cm$^{-3}$ has been corrected for an actual particle density of $\rho = 1.65$ g cm$^{-3}$. Similarly, *Rampino and Self* [1984] roughly estimated 10–20 Tg. Using lidar data and stratospheric transport modeling, *Cadle et al.* [1976] obtained 25 Tg.

Results derived by using methods not based on stratospheric optical thickness are presented in Table 4. In the case where precipitated sulfate has been measured in polar ice cores, most of the large uncertainly in $M_D$ is caused, for the Antarctic ice cores, by varying assumptions about the ratio of mass loadings between the northern and southern hemispheres and, for the Greenland ice core, by contamination from the Surtsey, Iceland, eruptions of 1963–1964 [*Hammer et al.*, 1980; *Legrand and Delmas*, 1987; *Delmas et al.*, 1992; *Cole-Dai and Mosley-Thompson*, 1999]. Other problems with the ice core method include the uncertain temperature-dependent polar ice chemistry, the locally variable deposition and erosion rates, and the globally variable long-range atmospheric circulation patterns. Methods based on the sulfur content of glass inclusions in the erupted magma and on the assumption of a fixed sulfur percentage of the total amount of erupted tephra yield results that are even less reliable. Although these results are also listed in Table 4, they will not be discussed further here, except to note that *Self and King* [1996] did upgrade their lower limit of 4 Tg to 14 Tg by employing additional considerations.

In the case of Fuego (1974) we derive from (5) a total aerosol mass of 4 Tg. Values in the range 2–5 Tg have been obtained from selected lidar measurements in conjunction with theoretical modeling of stratospheric transport [*Cadle et al.*, 1976, 1977; *Remsberg et al.*, 1982], while 2 Tg has been suggested by in situ sampling of aerosol particles [*Hofmann and Rosen*, 1977; *Lazrus et al.*, 1979]. Since our value lies near the threshold of what can be detected by the stellar extinction method, it does not fully settle the issue for Fuego.

**Table 4.** Previous Determinations of Volcanic Aerosol Loading of the Stratosphere During the Period 1961–1978

| Method | $M_D$, Tg Agung | $M_D$, Tg Fuego | Source |
|---|---|---|---|
| Turbidity | 15 | | *Deirmendjian* [1973] |
| | ~15 | | *Rampino and Self* [1984] |
| Tracer modeling[a] | 25 | 5.1 | *Cadle et al*. [1976, 1977] |
| | | 2.5 | *Remsberg et al*. [1982] |
| In situ sampling[a] | | 1.5 | *Hofmann and Rosen* [1977] |
| | | 2.1 | *Lazrus et al*. [1979] |
| Greenland ice[a] | 27 | | *Hammer et al*. [1980] |
| Antarctic ice[a] | 40 | | *Legrand and Delmas* [1987] |
| | 53 | | *Delmas et al*. [1992] |
| | 42 | | *Cole-Dai and Mosley-Thompson* [1999] |
| Glass inclusions[a] | | ≥0.9 | *Rose* [1977] |
| | ≥4 | | *Devine et al*. [1984] |
| | ≥4 | | *Self and King* [1996] |
| Total erupted tephra | 15 | | *Mitchell* [1970] |

[a]Published mass loading of $H_2SO_4$ multiplied by 1.33.

## 7. Discussion

Annual means of the hemispheric visual optical depth perturbations for about a century, 1881–1978, are listed in Table 5.

**Table 5.** Annual Mean Visual Optical Depth Perturbations (Multiplied by 1000) Due to Volcanic Eruptions During the Period 1881–1978[a]

| Year | Global | Northern Hemisphere | Southern Hemisphere |
|---|---|---|---|
| 1883 | 96 | 96 | 96 |
| 1884 | 192 | 192 | 192 |
| 1885 | 71 | 71 | 71 |
| 1886 | 26 | 26 | 26 |
| 1890 | 26 | 52 | 0 |
| 1891 | 48 | 96 | 0 |
| 1892 | 18 | 35 | 0 |
| 1902 | 14 | 16 | 12 |
| 1903 | 118 | 121 | 116 |
| 1904 | 61 | 61 | 61 |
| 1907 | 10 | 21 | 0 |
| 1908 | 4 | 7 | 0 |
| 1912 | 28 | 55 | 0 |
| 1913 | 19 | 39 | 0 |
| 1914 | 7 | 14 | 0 |
| 1922 | 8 | 0 | 17 |
| 1923 | 4 | 0 | 9 |
| 1928 | 2 | 0 | 5 |
| 1929 | 14 | 15 | 13 |
| 1932 | 12 | 0 | 23 |
| 1933 | 2 | 0 | 4 |
| 1962 | 12 | 0 | 24 |
| 1963 | 66 | 11 | 122 |
| 1964 | 51 | 26 | 76 |
| 1965 | 31 | 24 | 39 |
| 1966 | 14 | 18 | 10 |
| 1967 | 19 | 13 | 25 |
| 1968 | 11 | 8 | 14 |
| 1969 | 6 | 13 | 0 |
| 1970 | 6 | 13 | 0 |
| 1971 | 15 | 11 | 19 |
| 1972 | 8 | 9 | 7 |
| 1973 | 1 | 1 | 0 |
| 1974 | 7 | 11 | 4 |
| 1975 | 24 | 21 | 27 |
| 1976 | 7 | 11 | 4 |
| 1977 | 3 | 0 | 6 |
| 1978 | 1 | 0 | 3 |

[a]Missing entries are to be taken as zero.

These tabulated data have been compiled from our previous paper for the years 1881–1960 (with the necessary conversion from pyrheliometric effective wavelength to visual wavelength) and from the present Figure 4 for the years 1961–1978 (ignoring the uncertain South Pole data). They are plotted here in Figure 6, including an extension for more recent years, 1979–1994, taken from *Sato et al*. [1993] and from M. Sato and J. Hansen (personal communication, 1995).

A discussion of the whole time series, 1881–1994, was presented in our previous paper, although only rough data were available at that time for the period 1961–1978. Our main new conclusions, apart from the present provision of greatly refined data for this intermediate time period, pertain to a new assess-

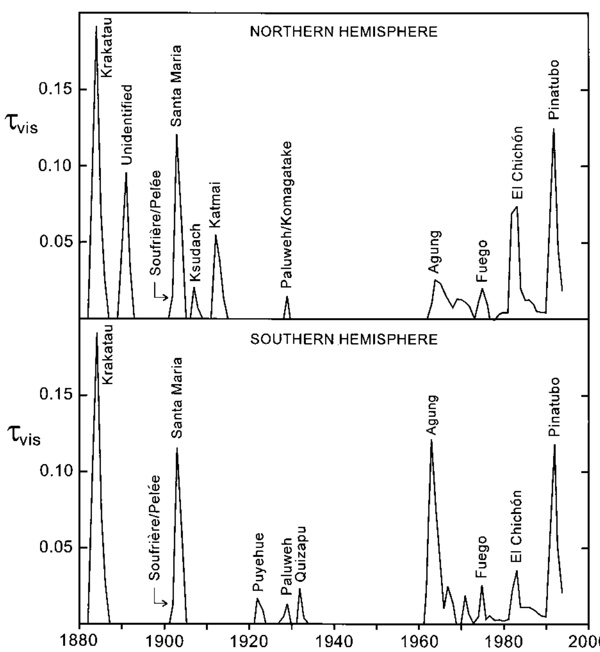


**Figure 6.** Mean annual optical depth perturbations at $\lambda = 0.55\ \mu$m for the northern and southern hemispheres. Data for the most recent years, 1979–1994, come from *Sato et al*. [1993] and M. Sato and J. Hansen (personal communication, 1995). Maxima are labeled with the source volcano.

ment of the Agung (1963) eruption. Agung's aerosol veil was the third most massive one of the twentieth century; only Santa Maria's (1902) and Pinatubo's (1991) exceeded it. Considering that this was a tropical eruption, a major peculiarity is that its aerosol veil spread so insignificantly into the opposite hemisphere. In the case of other large modern tropical eruptions, Tambora (1815), Krakatau (1883), Santa Maria (1902), Fuego (1974), and Pinatubo (1991), the mass loadings in the two hemispheres were approximately equal. Although in the case of El Chichón (1982) the hemisphere of the eruption remained more turbid by a ratio of about 3:1, this volcano lies somewhat farther away from the equator (17°N) than the others, and a ratio of 3:1 is still much smaller than Agung's 8:1.

These striking differences illustrate the complexity of trying to make predictions for the relative hemispheric mass loadings from volcanic eruptions without an accurate and detailed knowledge of the dynamical state of the atmosphere, especially the lower stratosphere. The data assembled in this paper, it is hoped, should aid in both checking and calibrating improved theoretical models of the general circulation of the atmosphere.

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